Terrestrial planet formation

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Advances in our understanding of terrestrial planet formation have come from a multidisciplinary approach. Studies of the ages and compositions of primitive meteorites with compositions similar to the Sun have helped to constrain the nature of the building blocks of planets. This information helps to guide numerical models for the three stages of planet formation from dust to planetesimals (~10^6 y), followed by planetesimals to embryos (lunar to Mars-sized objects; few × 10^6 y), and finally embryos to planets (10^7–10^8 y). Defining the role of turbulence in the early nebula is a key to understanding the growth of solids larger than meter size. The initiation of runaway growth of embryos from planetesimals ultimately leads to the growth of large terrestrial planets via large impacts. Dynamical models can produce inner Solar System configurations that closely resemble our Solar System, especially when the orbital effects of large planets (Jupiter and Saturn) and damping mechanisms, such as gas drag, are included. Experimental studies of terrestrial planet interiors provide additional constraints on the conditions of differentiation and, therefore, origin. A more complete understanding of terrestrial planet formation might be possible via a combination of chemical and physical modeling, as well as obtaining samples and new geophysical data from other planets (Venus, Mars, or Mercury) and asteroids.

How terrestrial planets grew out of the solar nebula has long been a topic of interest, and is still being pursued by a combination of studies involving samples of terrestrial and extraterrestrial materials, as well as computational models. Early nebular models by Descartes, Kant, and Laplace proposed that the planets formed from the solar nebula by various mechanisms such as vortices (analogous to spinning galaxies), or by shedding rings, but all of these general concepts failed to produce fundamental features of our Solar System (e.g., ref. 1). Subsequent models included homogeneous accretion (e.g., refs. 2, 3), in which the planets condensed from the nebula and were composed of homogeneous material that differentiated internally, and heterogeneous accretion (4), in which planets grew like layer cakes as the phases condensing out of the nebula changed with decreasing temperature. The purpose of this review is to elucidate how new samples, experiments, and modeling are contributing to a more thorough and sophisticated understanding of the many factors that helped to shape our inner Solar System (Fig. 1; see *SI Text*) and specifically the Earth. We also highlight areas where more extensive work is necessary or more samples are needed to better constrain our models, and where physical and chemical constraints may benefit from combined approaches.

Composition

Primitive chondrites were likely the building blocks for terrestrial planets, with bulk compositions similar to the Sun, but with differing oxygen fugacities (Fig. S1) and other chemical characteristics such as volatile element depletions (5; *SI Text*, Fig. S2). In addition to chondrites, differentiated small bodies, comets, and even unsampled bodies are other possibilities. Oxygen is the most abundant element in the Earth and other inner Solar System bodies, and the oxygen isotopic composition of a planet and its building blocks should be considered in detailed modelling. The variation in oxygen isotopic composition of chondritic materials is substantial, and must provide constraints on the kind of material that the Earth or other planets or asteroids were derived from (6; Fig. 2).

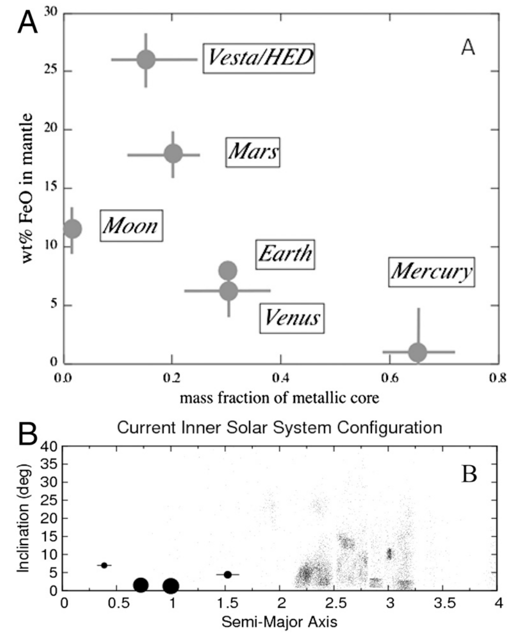

Fig. 1. (*A*): Mass fraction of metallic core and mantle FeO are roughly correlated for the terrestrial planets and an asteroid. A notable exception is Earth's Moon which has a very small core, presumably due to its unique origin by impact process. References for these properties are from (8). (*B*): Orbital structure of the current inner Solar System, relative to the invariant plane, illustrating the high eccentricities and inclinations of Mercury and Mars and much of the material in the asteroid belt. Horizontal bars show the range of peri/aphelion for the planets, but are not shown for the asteroids. Fig. 1 from *Meteorites and the Early Solar System, II*, edited by D. Lauretta and H.Y. McSween, Jr. Copyright 2006 The Arizona Board of Regents. Reprinted by permission of the University of Arizona Press.

Constraints on the composition of Earth come from thousands of direct samples studied extensively by the geological and geochemical communities (e.g., ref. 7), samples of meteorites which are possible building blocks for the planets (e.g., ref. 8), and seismic constraints from geophysical studies (e.g., ref. 9). Although these studies have led to a detailed understanding of the age, origin, and evolution of the Earth's crust, mantle, and core, a detailed bulk composition that satisfies all available geochemical data has remained elusive. For example, the general bulk composition (Mg, Al, Si) of Earth's primitive upper mantle is close to CV3 chondrites (Fig. S3), but there is not an exact match. The composition of the bulk Earth is commonly proposed to be defined by mixtures of various chondrite groups, other rare meteorite types, or even some chondritic material that doesn't exist anymore because it was all accreted by the Earth (10).

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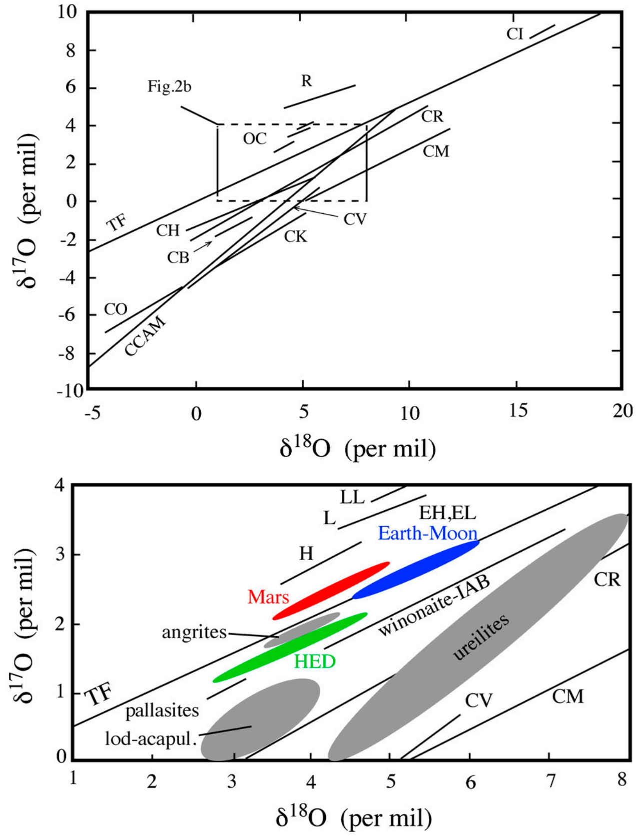

Fig. 2. (*A* and *B*): Simplified oxygen isotope plot of inner solar system material, based on the work of Clayton and colleagues, University of Chicago (e.g., ref. 6). TF line is the terrestrial fractionation line, and CCAM is carbonaceous chondrite-Allende mixing line. Also shown are fields or lines for the carbonaceous chondrites (CB, CH, CI, CK, CM, CO, CR, CV), ordinary chondrites (OC) and R chondrites, In *B* are shown fields for many differentiated bodies including Mars (martian meteorites), angrites, pallasites, lodranite-acapulcoites, ureilites, winonaites-IAB irons, and the howardite-eucrite-diogenite meteorites (HED). Ordinary chondrites are here split into the three subgroups LL, L, and H. Earth's Moon plots within the blue field along with terrestrial mantle samples and Enstatite chondrites (EH and EL). Mass dependent fractionation causes a slope of 0.5 on this diagram, whereas mass independent fractionation causes all other variation, such as the slope 1 : 1 line (e.g., caused by ^{16}O addition or subtraction).

Samples from the Moon (Apollo, Luna, and meteorites), as well as some lunar geophysical constraints, tell us about the differentiation history of the Moon (11). Furthermore, the link between the Earth and Moon through the giant impact theory has made aspects of Moon's geochemistry highly relevant to our understanding of Earth, such as their identical O isotopes (6). The composition and interior structure of Mars are constrained by studies of close to 60 martian meteorites, as well as multiple robotic spacecraft missions with operations both at the surface and in orbit (the latter constraining some key geophysical parameters such as core size, mantle density, and composition) (12). And we have nearly 1,000 meteorite samples (the HED clan; howardite-eucrite-diogenite) that are likely from 4 Vesta, a roughly 500 km diameter differentiated asteroid (13). Unfortunately we have no known samples from Mercury or Venus, and the few spacecraft missions to these planets have only provided minimal information regarding their bulk compositions and interior structures [e.g., MESSENGER mission providing chemical information on surface units (14), and the Venera robotic lander missions providing chemical analyses of surficial basalt; 15)]. Although basaltic materials (like martian meteorites) can provide information about mantle melting and interior composition, such can only be done with samples that are known to be fresh and unaltered. The martian meteorites include such samples, but robotic analyses of surficial samples on Venus (and some robotic Mars analyses) have no way of discerning fresh or altered material. This underscores a major advantage of sample return over robotic in situ analysis.

Mars can be made of a mixture of chondritic materials (e.g., ref. 16) and is thought to have accreted and differentiated within ~10 Ma of the start of the Solar System (17). The bulk composition, age, and formation history of Vesta, from which we have the HED meteorite samples, have been well studied and current ideas suggest it accreted from a mixture of known chondritic materials (CM + L chondrites), differentiated rapidly, within 2–3 Ma of T_0 (where T_0 refers to the time of first solid formation in the early Solar System—see below), and may have been substantially molten (e.g., refs. 18, 19). Vesta is not unlike the planetesimals or small embryos thought to have populated the early Solar System, with respect to its size, age, and thermal state, and therefore can provide constraints on planetary modeling scenarios (see next section).

Venus and Mercury are not well understood geochemically, except for some aspects of the Venusian atmosphere and their crusts. Aside from Mercury having a large core relative to its silicate mantle [perhaps due to a giant mantle-stripping impact; (20)] and FeO-poor surface (e.g., ref. 14), and Venus having an Earth-sized metallic core and Earth-like basalts at the surface (e.g., refs. 8, 15), detailed knowledge is not available yet.

As we collect more and more Solar System materials, new kinds of chondritic building blocks and differentiated materials are being found, indicating that there is extensive diversity even in the asteroid belt (21). Similarly, several years ago, a very unusual hornblende- and biotite-bearing chondrite was discovered from Antarctica, and it remains a one-of-a-kind water-bearing oxidized meteorite that could conceivably be a planetary building block (22). Unusual differentiated bodies have also been documented in the past decade with some reflecting conditions of differentiation not previously known, such as a more reduced version of the metal and silicate mixtures known as pallasites [e.g., the meteorite Itqiy; (23)], a still unknown parent body of the oxidized angrites (24), and an unusual feldspathic crustal meteorite that may represent a previously unknown style of planetesimal differentiation [GRA 06128, GRA 06129; (25)]. In summary, there is some uncertainty in the specific bulk compositions of planets because we do not know for sure what was available to be accreted onto the Earth, nor do we know how representative the sampled meteorites are of populations of material present from which the Earth formed.

Physical Models

The formation of terrestrial planets is generally divided into three major stages based on the different physical processes involved and their respective time scales (e.g., ref. 26): (*i*) dust aggregation into planetesimals; (*ii*) runaway and oligarchic growth of embryos from planetsimals; and finally (*iii*) formation of planets by high-velocity impacts between embryos (Fig. 3). All chronology is referenced to T_0, which is the abbreviation for the age of oldest known solid material in the solar nebula [e.g., ~4,568 Ma; (19, 27–29); see *SI Text*)].

Dust to Planetesimals (the Importance of Turbulence). The growth of planetesimals from dust grains in the nebula is not well understood, but a number of mechanisms have been proposed (see, e.g., ref. 30), including particle sticking, gravitational instability, turbulent concentration, and streaming instability. Rapid growth to sizes larger than 1 m is critical in any of these scenarios, as bodies around that size experience a strong gas drag effect that can drag them into the Sun on a time scale as short as ~100 y (31). This size threshold is often referred to as the "meter size barrier."

Because of the problems posed by turbulence for particle sticking and gravitational instability (see *SI Text*), a third class of models has been developed that explores growth in certain turbulent environments. For example, large ~10^3 km gravitationally bound

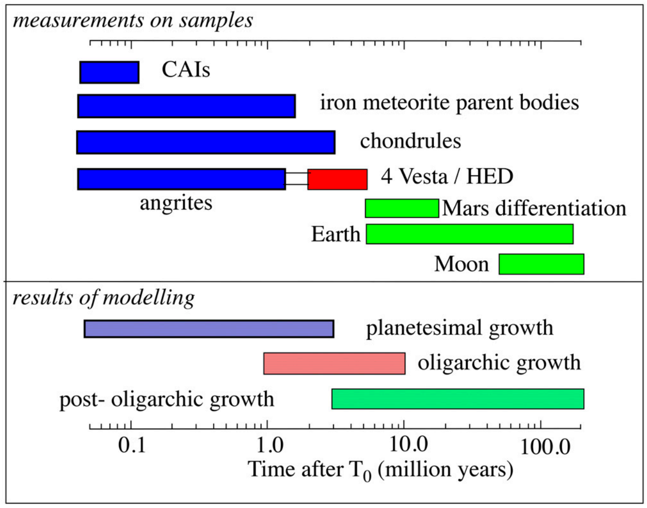

Fig. 3. Summary of the time scale of formation of various solar system objects from small scale CAIs and chondrules to differentiated planetesimals and embryos (such as angrites and HED parent body), to Mars, Earth, and Moon. The bottom half of the diagram shows the three different stages of planet growth that are discussed in the text, and the colors are keyed to measurements made on the real objects in the top. See text for references to individual samples and studies. Isotopic measurements are based on discussions in refs. 17, 19, 27–29.

clumps of particles can form in the stagnant regions between turbulent eddies via a process called *turbulent concentration* (30, 32). In these regions, high solid/gas ratios (~100) can occur, and gas drag and particle-particle collisions reduce the relative velocities of the particles in the clump. Under these conditions, bodies can grow rapidly beyond the meter size barrier (33). While the individual clumps may form and collapse rapidly, this process as a whole can occur over a time scale of a few million years, consistent with the range of ages recorded in components of meteorites (e.g., ref. 19).

Turbulence can also aid planetesimal formation in other ways: If a substantial fraction of solid material in the disk exists as meter-sized bodies, these bodies will rapidly drift towards and accumulate in the pressure maxima of turbulent regions (34). The clumps of bodies that form in these regions are protected from further inwards drift by the *streaming instability*, in which a clump of particles collectively moving through the nebular gas begins to drag the gas along with it, and thus experiences a reduced headwind. Other bodies drifting inwards from further out are added to these clumps, which can become gravitationally unstable, and as their relative velocities are damped by collisions, can grow into planetesimals as large as hundreds of km (35). Like the turbulent concentration scenario discussed previously, this scenario can also form individual planetesimals quite rapidly, but the process of planetesimal formation as a whole would occur for several Myr.

Even in scenarios where growth of individual planetesimals occurs rapidly, the planetesimals can form at a range of different times, capture material from different regions, and incorporate different components that formed at a range of different times, consistent with the range of materials and components found in meteorites and Stardust samples (e.g., refs. 36, 37). In addition, evaporation and condensation processes can operate on materials in the nebula which can then later be incorporated into larger bodies (21). Several processes can act to redistribute material in the Solar nebula, and potentially "store" early-formed material further out in the disk such that it can still be incorporated in meteorites that form later, such as radial diffusion in the disk, combined radial and vertical diffusion, or an x-wind scenario (38). Prolonged planetesimal accretion, and the potential for incorporating material of different ages, is consistent with measurements of ages of CAIs (calcium- and aluminum-rich inclusions) and chondrules ranging from 1 to 3 Ma after T_0 (39), and the ages of some planetesimal sized meteorite parent bodies [2 to 3 Ma after T_0; (18, 38)]. In addition, it is consistent with the time scales observed (Fig. S4) for primordial, gaseous protoplanetary disks around other stars (40), which have a median lifetime of ~3 Myr.

Planetesimals to Embryos (Runaway and Oligarchic Growth). Once planetesimals have formed and grow to a point where they can gravitationally perturb one-another, the process of *runaway accretion* can begin (e.g., ref. 41). The relative velocity in a swarm of planetesimals is governed by their frequent encounters with one another, and is kept low because of their small gravity. Because planetesimal relative velocities are dominated by their mutual interactions, their relative velocities are comparable to their escape velocities. Once a body starts to grow larger than the others, however, its geometric cross section increases and its gravitational field strengthens, allowing it to sweep up and accrete smaller planetesimals at an ever-increasing rate and begin the growth into a *planetary embryo*.

The transition from *runaway* to *oligarchic growth* occurs when the relative velocities of the planetesimals are stirred up by the growing embryos and start to become comparable to the escape velocity of the embryos (e.g., ref. 42). When an embryo begins to grow larger than its neighbors, it increases the relative velocity of planetesimals in its vicinity, which then decreases its accretion efficiency compared to other smaller embryos and prevents a single embryo from accreting the majority of the mass. The end result of this stage is several tens to ~100 Lunar- to Mars-mass embryos embedded in a swarm of remnant planetesimals.

Embryos to Planets (Giant Impact Stage). The final stage of terrestrial planet formation is characterized by high-velocity collisions among planetary embryos and remnant planetesimals over a span of ~10–100 Myr. Modeling of this phase is generally done with N-body simulations, with the goal of reproducing the total mass, mass distribution, eccentricity, and inclination of planets in our Solar System (43–51; *SI Text*).

With large numbers of bodies ($N > 1{,}000$), the more realistic situation of mixed embryos and planetesimals can be modeled, and the role of *dynamical friction* can be evaluated (e.g., refs. 48–50). Dynamical friction is the equipartition of energy between small and large bodies, which tends to decrease the orbital excitation of the largest bodies (i.e., the embryos and growing planets). The exploration of a wider range of initial configurations of Jupiter and Saturn has shown that they can exert a strong influence over the final mass distribution of terrestrial planets in the inner Solar System (e.g., refs. 52, 53), especially when the effects of secular resonance sweeping during the dissipation of the Solar nebula are taken into account (e.g., ref. 54). Strong turbulence in the nebula during the accretion of planets from a disk of embryos promotes the formation of four to five terrestrial planets with eccentricities and inclinations much like the planets in our own Solar System (55). Finally, if the planetesimals and embryos are initially confined to a narrow annulus in the terrestrial planet region, with an outer edge at ~1 AU, a system of terrestrial planets resembling the Solar System results, in particular forming a realistic Mars-mass planet (56). The disk of planetesimals and embryos could be truncated by the early gas-driven migration of Jupiter and Saturn in the protoplanetary disk (57). This scenario gives a good match to the terrestrial planets in the Solar System, and in addition the migration of the giant planets scatters material from different regions into the asteroid belt, explaining the diversity of spectral taxonomic types. Combinations of some or all of these five factors in modeling efforts have led to significant improvements in producing Solar-System-like terrestrial planets.

Several aspects of late-stage accretion deserve special note. First, the final stages of terrestrial planet formation involves mixing of material over significant radial distances (see Fig. 4). This mixing has important implications for water delivery and the chemistry of terrestrial planets, as discussed in more detail in the following sec-

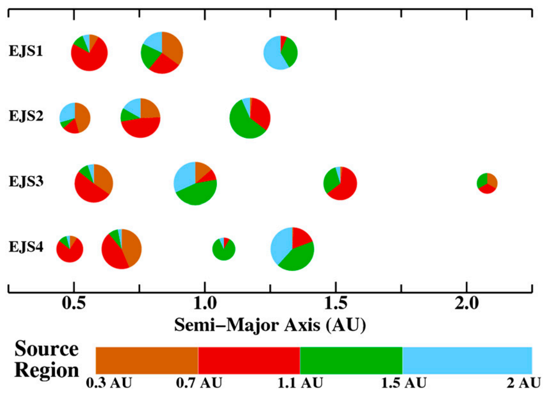

Fig. 4. The composition of the planets in a set of simulations from ref. 48, based on where the material in the planets originated (NOTE: material from beyond 2 AU is not included in this compositional breakdown). While it is clear that radial trends in composition are broadly preserved, there is significant stochastic variation, such that the location of a planet does not uniquely determine its composition.

tion. Second, the collision velocities between embryos in the final stages of planet accretion are large, and can lead to the ejection of significant amounts of material. For example, mantle stripping by a giant impact is the most likely reason that Mercury has a proportionately much larger core than the other terrestrial planets (20), and a giant impact is the likely origin of the Earth's moon (58).

Future and Unresolved

Bulk Composition of the Earth. There remains some uncertainty about Earth's bulk composition due to unknown compositions of deeper interior portions of Earth. There are three possibilities that can explain its bulk compositional properties. One is that it is made of known chondritic materials, but that the upper mantle has a different composition than the lower mantle. There is some isotopic evidence for this difference, and the geophysical data cannot rule this out, but it is unclear how a substantial distinct reservoir would survive 4.5 billion years of mantle convection. A second possibility is that it is made of known chondritic materials (enstatite chondrites), has a homogeneous upper and lower mantle composition, but the core contains Si (in addition to Fe, Ni, and other light elements), giving the mantle a higher Mg/Si ratio than chondrites. There is some chemical evidence for this possibility (59, 60) but the amount of Si indicated by the measurements is still debated. In addition, enstatite chondrites present other problems for a bulk Earth composition, and carbonaceous chondrites are generally favored over enstatite chondrites (e.g., ref. 61). A final possibility is that the Earth is made of chondritic material that is not currently represented in our meteorite collections (10). Ongoing research is aimed at resolving these uncertainties for the bulk Earth, given their fundamental implications for understanding terrestrial planets.

Origin of the Earth and Moon. The formation of the Earth-Moon system as a result of a giant impact has been the focus of significant modeling efforts (e.g., ref. 58). Using a computational approach called smoothed particle hydrodynamics, these models allow the mass, temperature, and density of material in two colliding bodies to be tracked, guided by equations of state for materials at high pressure and temperature conditions (e.g., ref. 58). As with other stages of the process, computational limits existed in early models, which typically modeled this stage with 100–1,000 particles (current models use 10,000 to 100,000 particles; 62, 63), and used an oversimplified equation of state. Although some aspects of modeling efforts remain problematic, unresolved, or unexplored [for example the role of water (62) and various ranges of dynamical conditions (63)], the simulations performed to date generally show that the Moon formed primarily from impactor material that was ejected into a postimpact circum-terrestrial disk (e.g., ref. 63). This result poses a challenge, because the similar oxygen isotopes for the Earth and Moon (6) suggest that the moon formed mostly from *target* material.

Geochemists lean to a connection between the Earth and Moon (or the Moon-forming impactor), whereas dynamicists argue that Earth and Moon-forming impactor were likely to have different compositions. The isotopic similarity between Earth and Moon for O, Cr, Ti, and other elements, as well as the results from modeling that suggest that most of the Moon is from the mantle of the impactor, lead to the geochemical inference that the impactor must have been basically the same composition as the Earth. This conclusion is often extended to imply that it must have formed right around where the Earth is (based on the earlier assumptions of narrow feeding zones). With the amount of radial mixing achieved in dynamical simulations, though, it would be essentially impossible for two bodies, even if formed at the same semimajor axis, to have the exact same bulk composition although some isotopic compositions may be the same (e.g., Fig. 4, Fig. S5). It is possible that there was some sort of postimpact equilibration (e.g., ref. 64), but there remain outstanding questions about these kinds of scenarios as well (65). Modeling that incorporates Hf-W isotopic data has frequently assumed complete isotopic equilibrium after impacts (e.g., ref. 66), but recent analysis of this problem suggests that there is only partial reequilibration of W isotopes after an impact, which has important implications for the modeling and interpretation of W isotopic anomalies (67). In summary, geochemical constraints imply a connection between Earth and Moon material, yet impact modeling efforts continue to yield Moons that originate from the impactor rather than the Earth, and this remains a conundrum.

Homogeneous or Heterogeneous Accretion? Two concepts have served as end member models for understanding planet and planetesimal formation for Earth, Mars, and 4 Vesta: homogeneous accretion and heterogeneous accretion. Both ideas were developed to explain geochemical features of the Earth, and were based heavily on partitioning of siderophile (iron metal-loving) elements between core and mantle (metal and silicate). Heterogeneous accretion held that Earth accreted reduced material early to form a core, and oxidized material later that formed the outer mantle (e.g., ref. 68). Homogeneous accretion held that Earth accreted as a homogeneous material, and then differentiated in response to internal and external heat sources (2). Both concepts have been used to explain aspects of Earth, Mars, and 4 Vesta; however successful models for all three bodies involve a mixture of both concepts (see *SI Text*; Fig. S6). Vesta and Mars can be explained by mixtures of known chondritic materials that differentiated into core and mantle (18, 69). For Earth there is not a known chondrite or chondrite mixture that satisfies its bulk composition, but mantle siderophile elements can be explained by high pressure and temperature metal-silicate partitioning at either fixed (70) or variable fO_2 (71). Ongoing studies should help resolve which model is more robust, and also provide constraints on the identity of the light element in Earth's core (S, C, O, Si, or H).

Combining Chemistry and Dynamics. Although still in the early stages, combining the physical modeling with chemical and compositional constraints has led to enhanced understanding of the planet formation process. For example, a late veneer of chondritic material is required in some geochemical models, and (48) demonstrated that it could be supplied in a realistic physical scenario. Water-rich embryos from the outer asteroid belt can contribute significantly to the overall water budget of any given terrestrial planet (72). Recent work (73) couples the results of a detailed nebular condensation model (which predicts initial planetesimal/embryo compositions) to an N-body model of planet formation. It is expected that future efforts to combine these two fields will lead to even more discoveries and a better under-

standing, especially when applied to terrestrial planet formation in extrasolar planetary systems which may have significantly different C/O and Mg/Si ratios than the Solar System, and hence very different solid materials available for planet-building (74).

Another aspect of the modeling that has led to breakthroughs in understanding the planet formation process is the degree of radial mixing in the early Solar System. Although early models (43) indicate limited mixing and narrow feeding zones, more accurate N-body modeling shows that radial mixing becomes more extensive over time, with some final planets receiving significant amounts of material from $\ll 1$ AU out to several AU (Fig. 4; 26, 48, 56). Many of the primitive C-type asteroids in the main asteroid belt may have been implanted there from beyond ~5 AU (57). As we obtain more information from Solar System materials regarding compositional or isotopic zoning, the interaction between chemistry and dynamics will be more fruitful. For example, isotopes of the refractory element Ca exhibit no variation for many differentiated bodies in the inner Solar System (75). Similarly, the volatile element *N* exhibits a narrow range in isotopic composition for all known planetary materials, yet this isotopic composition is distinct from the compositions of the Sun and comets (76). How did these isotopic ratios become homogenized (Fig. S7) when there are clear differences in the isotopic ratios of other elements such as O (6) (Fig. S8)?

One final area where the interaction of dynamics and chemistry needs to be treated more thoroughly is in the general field of giant impacts, which are common during the late stages of terrestrial planet formation and may have important implications for their current compositions. Although it is recognized that high temperature condensation from a gas after a giant impact will control the chemistry of solids and liquids (e.g., high temperatures in models of ref. 58), application of such a model to the postimpact Moon has not yet been carried out. Combining chemistry and dynamics together will be difficult because, for example, the condensation of solids leads to opacity changes, which change how the proto-lunar disk cools and evolves. However, coupling of impact and condensation models remains a worthy goal and may lead to a better understanding of compositional aspects of the Earth-Moon system. In addition, the loss of volatile elements during embryo collisions may lead to significant deviations from chondritic compositions for a large subset of elements (77). Indeed impact erosion during accretion and planetary growth may be common and can affect the overall chemistry of the bodies involved (78).

Habitability. One of the primary features of a habitable planet is the presence of water (79). Proposed sources of Earth's water have included comets, water from planetesimals and embryos, and nebular gas (79; see *SI Text*). Another point relating to habitability is the role of giant impacts in forming satellites around terrestrial planets. It has been shown that the presence of the Moon has had a stabilizing influence on the rotational dynamics of the Earth, thus promoting a habitable environment for life (80). Although the giant impact hypothesis is a front-runner for origin hypotheses for the Earth-Moon system, there are some aspects of this hypothesis that are not well understood, such as the importance of vaporization, impact angle, and the degree of postimpact mixing between the Earth and lunar disk (e.g., refs. 63–65). The absence of a satellite around Venus (e.g., ref. 81), coupled with its unusual rotational characteristics have led some to conclude that Venus experienced two giant impacts late in its history (82). Aspects of the giant impacts should continue to be tested to allow a more thorough understanding of the role of satellites in promoting a habitable environment on terrestrial planets.

Additional Samples or Observations. Although considerable progress has been made in recent years regarding planet formation, some important information that would greatly aid the understanding of our own Solar System is still missing.

The bulk and isotopic composition of Mercury and Venus remain poorly known. It is currently not possible to evaluate terrestrial planet formation in our Solar System for these two planets the same way we can for Earth and Mars. For example, it is possible that the oxygen isotopic composition of Mercury and Venus is the same as for Earth, in which case the fact of the Earth and Moon having identical O isotopic composition becomes less significant. On the other hand, if all three planets have different O isotopes, then that knowledge would help us to evaluate models of planet formation, namely whether or not particular models are able to preserve such heterogeneities. Having samples of these planets would also allow their differentiation ages and the heterogeneity or homogeneity of isotopic systems to be evaluated and integrated into the sophisticated framework of understanding that exists for Earth and Mars. Sample return from Mars could also have a potentially huge impact on our understanding of planetary accretion —differences between the martian meteorite collection and compositional data collected by the Mars Exploration Rovers and Pathfinder (11) illustrate that our small sampling of this geologically diverse planet limits a truly global understanding.

Direct links between meteorite samples and specific bodies or groups of bodies in the asteroid belt remain limited to asteroid 4 Vesta and the HED meteorites (e.g., ref. 12). Collection and return of samples from known and well characterized asteroids would greatly enhance the value of the enormous number of samples we have in our meteorite collections. Such connections would ultimately influence and improve our understanding of planetesimal and embryo formation and therefore terrestrial planet formation.

Secondly, we have only a general understanding of the variability of H_2/H_2O ratios in stellar systems (83), but this ratio can control not only the bulk compositions and metal content of chondrules and nebular condensates, but also ultimately the sizes of planetary cores. For example, is the Earth's moderate sized metallic core and FeO-bearing mantle typical for terrestrial planet formation? Or is it unusual? The controls on H_2/H_2O and oxygen fugacity in general in the early nebula as well as later stages of growth are critical to understand because they affect the amount and distribution of oxidized and reduced Fe that is available for planet making (84). Will this be an issue in other solar systems, or is this something specific to our own?

Furthermore, resolving the origin of terrestrial planet systems will benefit from observations of other planetary systems, especially those containing "super-Earth" planets, a number of which are currently known, and planets comparable in mass to the Earth, as are expected to be discovered by the Kepler mission. These new discoveries will no doubt aid our understanding of habitable Earth-like planets (e.g., refs. 49, 50).

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